Analysis of ALOHA-93 Campaign Data in Terms of Gravity and Tidal Wave Modes: Considerations on the Jet Stream as a Gravity-Wave Source

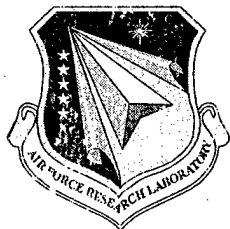T.F. Tuan

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14. ABSTRACT

We report on two particular phenomena observed in the ALOHA-93 Campaign. On 20 Oct 1993, the data showed a sudden sharp rise in temperature over an exceedingly narrow height range, with an initially modest temperature peak at 93 km rising to a peak value two h later at a height of 88 km, then subsiding to a much smaller value near 87 km some 4-1/2 h later. The second phenomenon is the simultaneous obseraation of OH airglow wave structure propagating along an azimuth of 340 deg with a phase speed of 35 m/s, a horizontal wavelength of 80 km remaining for the duration of temperature-inversion layer observation.

Our analysis shows that a critical layer exists at about 87 km between the beginning of observation at 0830 UYT until after 1200 UT. By plotting the vertical gradient of horizontal wind we show that it has a maximum in the 340-160 deg direction. By plotting the Richardson number vs height, we show that the shear grows and becomes sufficiently large to initiate the KH instability only over a height range of 2 to 3 km centered at 87-88 km between 1000 and 1230 UT. We conclude that the source of energy and momentum for the temperature inversion layer and associated large wind shear observed near 88 km is a downward-propagaiing gravity wave interacting with the total background wind field, including the diurnal tide.

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SUMMARY

In this summary we will report on two particular phenomena observed in the ALOHA-93 Campaign. 1) The first phenomena occured on October 20th, 1993 (day 294) when the data showed a sudden sharp rise in temperature over an exceedingly narrow height range with its initially (8:30 UT) modest temperature peak at a height of 93 KM, rising to its maximum peak value two hours later (10:30 UT) at a height of 88 km before subsiding to a much smaller value at about 87 km some four and half hours later (13:00 UT), (Gardner, 1995, Tao and Gardner, 1995, and private communication, 1997). 2) The second phenomenon is the simultaneous observation (Swenson et al, 1995, and private communication, 1997) of OH airglow wave structures propagating along an azimuth of 340 degrees (i.e. 20 degrees west of north, or NNW). The wave had an observed phase speed of 35 m/s, a horizontal wavelength of 80 km, and remained for the duration of the temperature-inversion layer observation.

At the suggestion of Dr. E. M. Dewan of AFRL and also from the observations of Dao et al (1995) which show rather fine small-scale wave structures with vertical wavelengths of the order of 8-9 km above a height level ranging from 85 to 90 km superimposed over a large tidal wave mode with vertical wavelengths of the order of 20 km, we have decided to adopt a model for the inversion layer in which a high-altitude gravity wave source from a possible equatorial electrojet at about 100 km produces a gravity wave that reaches Hawaii with a downward propagating component, (downward in energy propagation). The wave then interacts with the horizontal winds at about 87 km at a critical layer below which there would be no small-scale gravity wave. The sudden temperature rise over such a narrow localized altitude range (about 3-4 km in half-width) could then be attributed to the heat energy generated from the very thin layer (again, about 3-4 km) of turbulence from the gravity wave-critical layer interaction. Indeed, although at the beginning of the observation there is already a gravity wave and a critical layer, the layer vanishes after 12:00 UT (owing to a drop in wind speed to below 35 m/s) and the temperature also drops, thus reinforcing the critical-layer theory for the temperature-inversion layer.

(1)

We divide our investigation into two parts: the first part deals with a purely phenomenology investigation of the relationships between different types of simultaneous observations. We have plotted the Richardson number for the background wind profile at all the different times. The results show: (1) the Richardson number drops very sharply over a very narrow vertical range near the critical layer; (2) from 10-11:30 UT the Richardson number dips below ¼ at about 87-89 km when instability in the time-averaged background mean flow would begin to occur. We wish to point out that even at 8:30 UT the oscillating gravity wave would already be unstable and hence some heating must occur since there is then already a critical layer, but the time-averaged background wind has not yet had time to acquire enough energy-momentum transfer from the gravity wave to be unstable. After 12:00 UT the wind speed drops to below the horizontal phase velocity of the wave and the critical layer vanishes, removing the source for turbulence and heat production. This can account for the observed drop in the temperature peak after 12:00 UT.

We then plot at different height levels the observed vertical wind gradient along different possible wave propagation directions over a 360 degree range in azimuth. The results show two peaks in the wind gradient: one at 160 degrees and one at 340 degrees. Since the observed wave propagation direction is along 340 degrees and critical layer theory tells us that the maximum energy-momentum transfer occurs along the direction of wave propagation, one would expect the wind gradient (corresponding to maximum energy-momentum transfer) would be the greatest at 340 degrees and, of course, its opposite direction at 160 degrees. Furthermore, the maximum energy-momentum transfer corresponding to the largest peaks in the wind gradient occur at 87 km or very close to the critical layer where the instability and turbulence created by the wave-mean flow interactions is maximum.

The results of this part of the research are published in J. Geophy's Res., 103, 6323-6332, (1998) under the title "Sudden narrow temperature-inversion-layer formation in ALOHA-93 as a critical-layer interaction phenomenon".

(2)

The second part of our report provides a more quantitative analysis of the gravity wave-critical layer model. Assuming a horizontally uniform but otherwise arbitrary atmosphere, we use the coupled linearized hydrodynamic equations in the presence of horizontal winds and eliminate the dependent variables such as velocity and density to obtain a single second-order Sturm-Liouville differential equation for the pressure perturbation. The solutions for the other dependent variables such as velocity and density, etc, can be readily computed once we have the pressure perturbation. The equation is similar to but more general than the Schroedinger equation.

In deriving the equation for the modification of the mean flow by the gravity wave, Lindzen (equation (8.27), Lindzen (1990)) had to introduce implicitly some small amount of wave damping, or absorption. This would also allow him to write down equation (8.28) where he assumed that the energy absorbed from the wave goes into (1) the rate of change of kinetic energy of the mean flow and (2) the rate of change in the internal heating. This formulation, however, does not explicitly consider the effects of the wave absorption on the gravity wave itself that means that the gravity-wave solutions used in the above equations would not contain the effects of damping which may be the reason that the damping had to be assumed small.

Since, in fact, the damping and dissipation are very large, and the gravity-wave solutions are severely affected, we should begin by explicitly introducing a model for gravity-wave damping in the gravity-wave equation. We refer here to the "optical model" method which works well for large damping and dissipation and is used whenever we need to determine the behavior of what remains of the original wave when it has lost most of its energy and momentum through different physical mechanisms. The method has been used in nuclear physics, for instance, when we wish to determine the behavior of the elastic scattering processes in the presence of inelastic excitations and/or ionizations all of which may be simulated by absorption of the original incident wave through introducing an imaginary "optical" potential in the Schroedinger equation. The Schroedinger equation is then no longer Sturm-Liouville, i.e. no longer self-adjoint, and the wave flux no longer be conserved, thus allowing the original incident wave to lose energy to the inelastic processes.

Since, for our case, the incident gravity wave from high altitudes suffers a very large dissipative loss in energy due to its interaction with the critical layer and above all we are primarily interested in the damped gravity-wave solutions, not in the explicit details of how the losses occur, our gravity-wave equation is very suitable for the optical model treatment.

We do know from our physical model in part (1) however, that all the energy lost within the narrow critical layer region is ultimately converted into (1) the change in the kinetic energy of the mean flow; (2) the increase in internal heat energy of the mean flow which serves to produce the sharp temperature-inversion-layer. Thus, using the modified Sturm-Liouville equation mentioned above, we can derive an energy-flux equation for the vertical energy transport and the dissipation energy density in terms of the optical potential. Here, we are fortunate to be dealing with dissipation within a vertical range (~3 km) that is very narrow compared to the vertical wavelength of the incident wave (8-10 km). Thus, similar to low-energy nuclear physics, where a two-parameter potential is required (Bethe, (1949); Bethe and Morrison, (1956); Austern, (1969); and Satchler, (1983). Actually, we only use one parameter which is the depth of the optical potential, W degrees since the effective range can be determined by the width of the sharp drop in the Richardson number (computed from observations) near the critical layer where gravity-wave absorption and therefore heating can occur (Huang et al, 1998). To take care of the rapid increase in absorption as we approach the critical level itself we assume that the absorptive optical potential is inversely proportional to the Doppler-shifted frequency.

Since the observed wind and temperature profiles over a height range 85-100 km were time-averaged over half-hour intervals, beginning at 8:30 UT and ending at 13L00 UT, we will use the observed wind profile as imput wind for each half-hour time slot and solve our gravity-wave equation for the particular time slot to obtain, over the same height range, the wave solutions, which can then be inserted into the energy-flux equation to obtain the rate of energy density dissipation by the wave.

We should point out that since the widths of the valleys in the Richardson number vary with different half-hour intervals, the range of our Optical Potential is time dependent in the sense that they very with each time slot in the same way that it varies with each time slot in the same way that the input wind profiles. Furthermore, the observed wind profiles would of course include the downward tidal wave motion (or any other large-scale background motion) as well as the energy-momentum transfer by the gravity wave.

Our principal assumption then is that the rate of energy density loss from the wave is equal to the rate of increase of the kinetic energy density and internal heat energy density of the mean flow, which can then replace the energy density dissipation term in the energy-flux equation just mentioned to obtain Lindzen's equation (Lindzen, 1990). Our Optical Potential can than be considered as an explicit model for the "damping" assumed by Lindzen (1990). Using the usual assumption that the rate of change of the momentum density of the mean flow is given by the rate of deposition of the horizontal momentum density by the gravity wave, we can simplify the Lindzen equation to obtain a simple equation for the heating rate (after first expressing the pressure perturbation in terms of the temperature perturbation assuming the usual adiabatic wave compression and the perfect gas law).

The temperature can then be obtained through the numerical integration of the heating rate equation after adding a cooling parameter, C, (which forms our second parameter) and using the gravity-wave solutions we have already obtained for each time slot. There is no problem with either numerical instability or accumulated errors because we use the "exact" input (i.e. the observed wind profile) for each successive time slot. Had we to compute a new wind profile from the energy-momentum transfer computed in the previous time slot for each succeeding time slot, we would have been closer to a nonlinear treatment, but both numerical instability and accumulated errors would then have occurred due in part to the length of the time slot (half an hour).

The computed results for the temperature profiles for every one of the half-hour time slots show very good agreement with the observed temperature profiles using onlt two fixed parameters, W and C, for all the different time slots. A plot of the rate of momentum deposition against the background Richardson number shows that: (1) it is large close to the critical level and within the widths of the sharp drops in the Richardson number (when the background is least stable); (2) it is large at earlier times when only the gravity wave is unstable, and the averaged background winds have not yet received enough momentum transfer to be unstable; (3) it vanishes after 12:00 UT when the critical level has vanished. We also find that the rate of change of temperature follows the rate change of momentum deposition. In other words, a greater rate of momentum deposition brings a smaller temperature change all seems to make good physical sense.